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ANNUAL REPORT FOR THE YEAR

August 1, 1978 - July 31, 1979

STUDIES OF RADIATIVE TRANSFER IN THE EARTH'S ATMOSPHERE WITH EMPHASIS ON THE INFERENCE OF THE RADIATION BUDGET

IN THE

JOINT INSTITUTE FOR ADVANCEMENT OF FLIGHT SCIENCES

AT THE

NASA-LANGLEY RESEARCH CENTER

NASA Grant NSG 1555

(NASA-CR-158690) STUDIES OF RADIATIVE TRANSFER IN THE EARTH'S ATMOSPHERE WITH EMPHASIS ON THE INFLUENCE OF THE RADIATION BUDGET IN THE JOINT INSTITUTE FOR ADVANCEMENT OF FLIGHT (George Washington

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School of Engineering and Applied Science
The George Washington University
Washington, D. C. 20052

During the time since the initiation of this grant, the major activity in support of this grant has been Dr. B. R. Barkstrom's work on the Ad Hoc Committee on the Earth Radiation Budget Measurement Process, with emphasis on sensor calibration and measurement accuracy. Dr. Barkstrom's contribution to this committee involved a significant fraction of his time since the inception of the committee on November 7, 1978. The work included the preparation of summaries of the past work on the earth's radiation field that must be used in reducing observations of the radiation field. Preliminary portions of the committee's final report are included here as appendices, and represent the formal portion of Dr. Barkstrom's contribution to this committee.

Dr. Barkstrom also provided significant contributions to the research in the Atmospheric and Environmental Sciences Division by aiding with the implementation of the finite difference radiative transfer algorithm for development of models of the angular and spectral dependence of the earth's radiation field for use in data reduction of the Nimbus 6 scanner data. He has also been developing a simple version of the optimization algorithm.

APPENDIX A

TASK 1

REVIEW OF EARTH AND SOLAR RADIATION ESTIMATES

Energy Balance

The simplest possible model of the Earth would be a sphere of constant and uniform albedo, isolated in outer space, illuminated by the sun, and emitting its own thermal energy into the void. In a steady-state condition, the energy (in W) accepted from the sun (given by the left side of equation 1) is balanced by the emitted thermal energy (given by the right side of equation 1); that is,

$$\pi R_{\oplus}^2 (1 - \bar{a}) F_{\odot} = 4\pi R_{\oplus}^2 F_{\oplus} \qquad . \qquad (1)$$

In this equation, there are four unknowns: $R_{\oplus}$, the radius of the Earth; $\bar{a}$, the global albedo of the Earth; $F_{\odot}$, the solar constant; and $F_{\oplus}$, the emitted (infrared) flux.

The (mean) radius of the Earth $R_{\oplus}$ is known to be 6371 km from geodetic surveys. A generally accepted value for the extraterrestrial solar ~~flux~~ irradiance $F_{\odot}$ ~~density~~ is 1360 Wm^{-2} (or 2 cal cm^{-2}). The global albedo $\bar{a}$ is generally thought to be around 1/3. It would then follow from equation (1) that the (average) emitted thermal energy density is $F_{\oplus} = 227\ Wm^{-2}$.

That an energy balance exists appears to be a good assumption since life has existed on the surface of the Earth for a long period and only small changes in the energy balance would cause drastic changes in climates.

However, estimates of global albedo and emitted thermal energy are ~~admittedly~~ uncertain.

Winds and Climates

Even before a quantitative statement of global energy balance was possible in the closing decade of the nineteenth century, the experiences of the European explorers led natural philosophers and scientists to speculate on the causes of the winds and climates of different regions of the Earth. Hadley in a famous paper in 1735 seems to have been the first to suggest an explanation of the large scale features that is still most commonly accepted. He suggested that solar heating near the equator would force air upwards and that the cooling near the poles would force air downwards. By insisting upon conservation of angular momentum, the tradewinds could be explained.

The important conclusion from this explanation is that atmospheric circulation is caused by imbalances between incident solar energy and the emitted thermal energy. The gross features of this imbalance are shown in figures 1 and 2 (from Lorentz, 1967). The latter figure illustrates how the northward or southward transport of energy depends on the latitudinal imbalance between the solar input and the terrestrial output.

Atmospheric circulation is, of course, much more complex than northward or southward movement. The distribution of land and water is exceedingly uneven, and their radiation properties differ markedly. Thus the reflected and emitted energy vary, producing marked variations in the

energy balance of the Earth's surface, as illustrated in figure 3 (from Budyko, 1978).

Although figure 3 shows an average amount of energy available to be converted to other forms, it does not show the expected variations (with time) in the energy driving both atmospheric and oceanic circulation. Nor does it give any indication of energy available to evaporate water, which may form clouds and precipitate elsewhere.

Effect of Changes in Energy Balance

A decrease in the incident solar irradiance of 1% (about 14 Wm^{-2}) would result in a 1.5°C decrease in the mean surface temperature of the Earth if cloudiness and earth albedo remain constant. If this small decrease would become permanent, then the border of the polar ice cover would shift from 10° to 18° in latitude toward the equator, and the mean surface temperature of the Earth would decrease by 5°C. These changes correspond roughly to those that occured during the quaternary glaciations [SMIC, 1971].

A change in the earth albedo of 0.01 (e.g., from 0.30 to 0.29 or about 14 Wm^{-2} in the reflected radiant exitance) would lead to a 2.3°C change in the mean surface temperature due to redistribution of radiative sources and sinks [SMIC, 1971]. Problems resulting from the sensitivity of Earth conditions to small changes in the energy budget are treated in numerous publications, including the study Inadvertant Climate Modification (SMIC, 1971) from which these illustrations are taken.

These types of concerns have led to the summary given in Table 1-1 as taken from the report Toward an Internationally Coordinated Earth Radiation

Budget Satellite Observing System: Scientific Uses and Systems Considerations (1978). (How about a short summary of Table 1-1 relevant to ERBE specifications?) ~~measurements?)~~

A Short History of Solar Radiation Estimates

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A Short History of Earth Albedo and Radiation Estimates

Before the advent of satellites, there were two major methods of estimating the Earth albedo. Figure 4 ~~Table 1-2~~ summarizes some of these estimates as well as those obtained from satellite measurements. Most material of the following discussion is taken from Levine [1967].

(1) In the meteorological method, the reflected solar irradiance is ~~was~~ assumed to be made up of three components: ground reflection, atmospheric reflection, and cloud reflection. By developing estimates of the fraction of the earth covered by clouds and their reflectivity, the global albedo could be found.

That is, the reflected radiation is assumed to be the sum of three components:

$$\bar{a}\, F_{\odot} = [f\, a_c + (1-f)\, (a_e + a_a)]\, F_{\odot}$$

where $\bar{a}$ is the earth global albedo, $F_{\odot}$ is the incident solar irradiance (about 1360 ~~1400~~ Wm^{-2}), f is the fraction of the earth covered by clouds, a_c is the cloud albedo, a_e is the albedo of the earth, and a_a is the albedo of the atmosphere. The most widely accepted estimates were given by London in 1957: for a year's average, $f \approx 0.51$, $a_c \approx 0.47$, $a_e \approx 0.139$, and $a_e \approx 0.086$. Thus, $\bar{a} \approx 0.35$, and 2/3 of the reflected solar radiation is due to clouds.

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(2) Another method of estimating the Earth albedo is to ~~by~~ measure ~~measuring~~ the earthshine reflected from the moon. This ~~type of~~ procedure has been used

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by Danjon (1936, 1954) and Bakos (1964). Generally, the earthshine measurements produce a larger albedo than that suggested by other recent measurements. Furthermore, the earthshine measurements suggest that the albedo is highest in the Northern hemisphere winter, whereas meteorological methods suggest that it is lowest then.

The method has been criticized on a number of grounds, including (1) that the observations so far do not include large areas of the earth (since the observations used come mainly from France and thereby exclude the Pacific), (2) that the angular dependence of reflection upon time of year has not been properly accounted for, and (3) that the measurement excludes the near infrared region of the spectrum.

Clearly, the major difficulty in developing accurate estimates of the reflected component of the earth radiation budget is due to the uncertainties in the area of the earth's surface covered by cloud and in the reflectivity of the clouds themselves. For example, Robinson (1956) has estimated the reflectivity of an overcast sky to be 0.65, while Neiburger (1949) derived a value of 0.75, and Roach's aircraft measurement (1961) gave a value of 0.85. This spread is characteristic of both observations and theoretical estimates.

(3) Discussion of spacecraft measurement results up to Nimbus 6.

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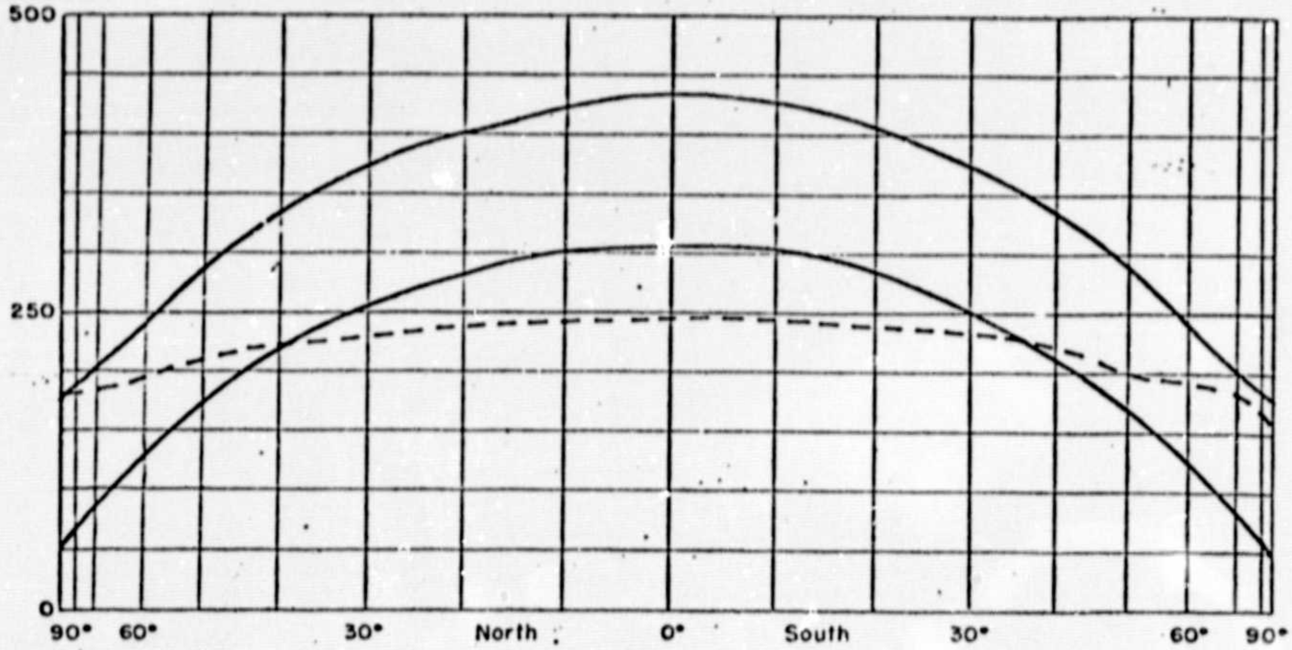

Figure 1 — Average solar energy reaching the extremity of the atmosphere (upper solid curve), average solar energy absorbed by the atmosphere-ocean-Earth system (lower solid curve), and average infra-red radiation leaving the atmosphere-ocean-Earth system (dashed curve), as given by Sellers (1968). Values are in watts m^{-2} (scale on left). (1 watt m^{-2} = 1.435×10^{-3} cal cm^{-2} min^{-1} = 0.754 kilolangleys $year^{-1}$.)

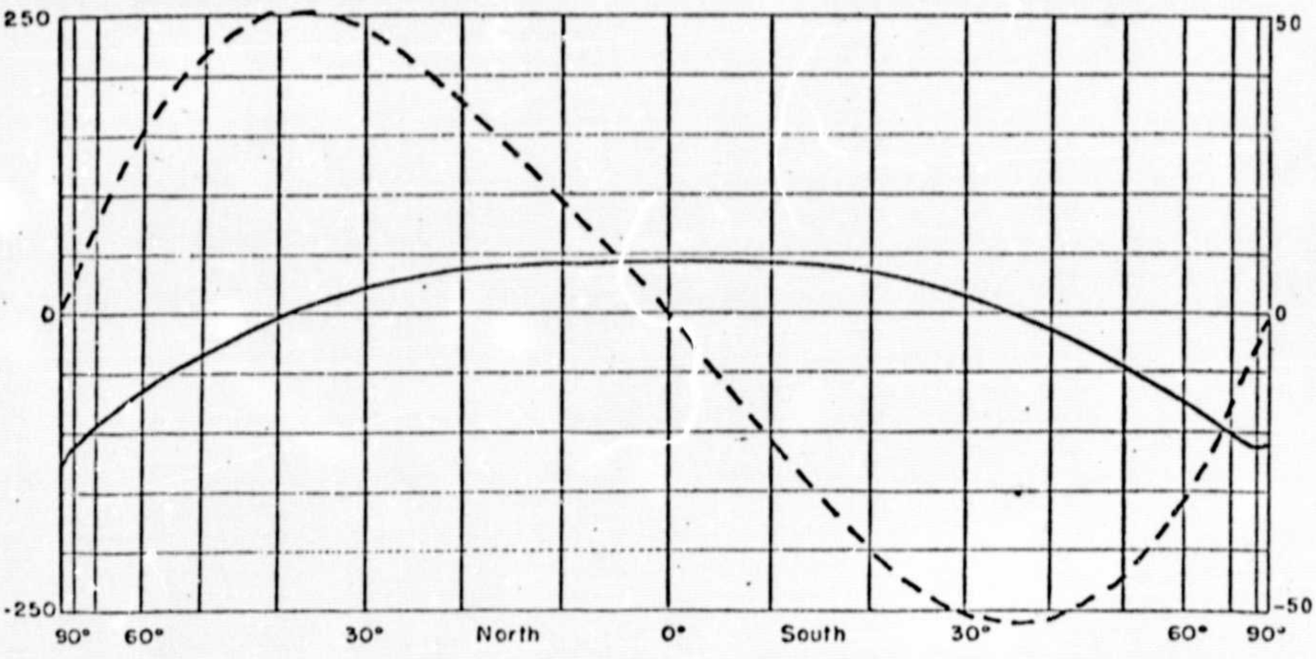

Figure 2 — Excess of absorbed solar radiation over outgoing infra-red radiation (solid curve), as given by Sellers (1966), in watts m^{-2} (scale on left); and northward transport of energy by the atmosphere and oceans required for balance (dashed curve), in units of 10^{11} watts (scale on right)

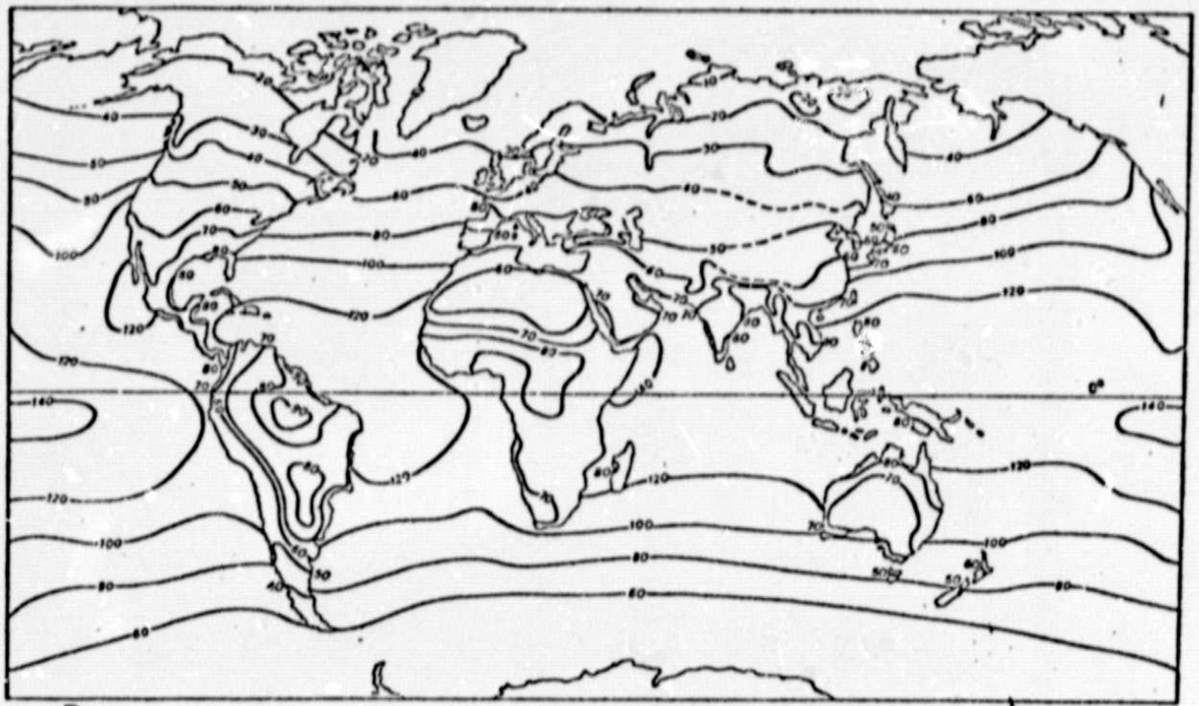

Fig. 3 The radiative balance of the Earth's surface (kcal $cm^{-2}yr^{-1}$).

10 kcal cm^{-2} yr^{-2} = 13.3 Wm^{-2}

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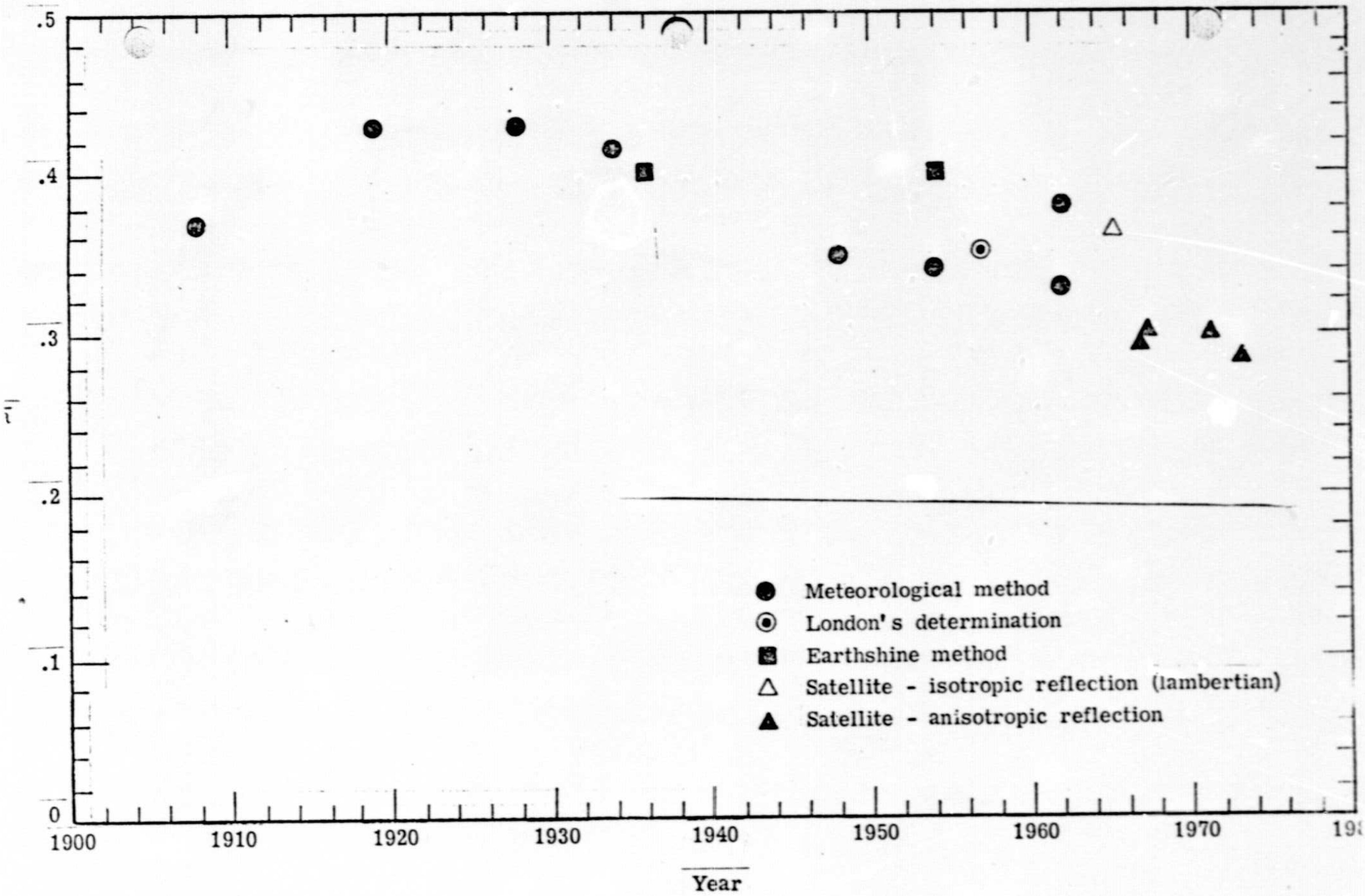

Figure 1-4. Estimates of Earth global albedo ($\bar{a}$).

TABLE 2. THE USE OF EARTH RADIATION BUDGET DATA FOR STUDIES OF CLIMATE (THE DETAILED RATIONALES ARE DISCUSSED IN RELEVANT PORTIONS OF THE TEXT)

AREA	USE	QUANTITIES TO BE MEASURED	SYSTEM ACCURACY	SCALE SPACE/TIME	CONSTRAINTS, COVERAGE
1. Climate diagnostics	atmospheric energetics	total net radiation at top of atmosphere and at earths surface	$\pm$ 20 watts m^{-2}	$\leq$ 2000 km, 1 mon.	global
	ocean heat transport	total & net radiation for period of 5-10 years	$\pm$ 5 watts m^{-2}	1000-2000 km regions, 10° lat. zones; 1 month	global
	surface radiation budget	total net radiation	$\pm$ 20 watts m^{-2}	200-1000 km regions; 1 month	global
2. Climate monitoring	variability of solar irradiance	total solar irradiance; spectral components. BOTH FROM WHOLE DISC	accuracy 0.5% * precision 0.3%*	monthly, over at least one solar cycle	must observe entire solar disc
	documentation of climate change	total irradiance, spectral comp'nts	accuracy 0.5% $<$ 1%, depending on spectral range	monthly daily	must observe entire solar disc
		net radiation at top of atmosphere			
		global	$\pm$ 2 watts m^{-2}	monthly	global
		zonal	$\pm$ 10 watts m^{-2}	10° 1 month	global
		regional	$\pm$ 20 watts m^{-2}	1000 km;1 month	global
		reflec. solar rad			
		global	$\pm$ 2 watts m^{-2}	1 month	global
		zonal	$\pm$ 10 watts m^{-2}	10° L&L;1 month	global
		regional	$\pm$ 20 watts m^{-2}	1000 km;1 month	global
		sfce rad. budget			
		global	$\pm$ 2 watts m^{-2}	1 month	global
		zonal	$\pm$ 10 watts m^{-2}	10° L&L;1 month	global
		regional	$\pm$ 20 watts m^{-2}	1000 km;1 month	global
		surface albedo			
		global	0.005	1 month	global
		zonal	0.06	10° L&L;1 month	global
		regional	0.03	1000 km;1 month	
3. Climate theory	validation of climate models	reflec. solar rad	$\pm$ 5 watts m^{-2}	500 km; 1 month	global
		emit. LW radiat'n	$\pm$ 5 watts m^{-2}	500 km; 1 month	global
		clear sky albedo	0.02	500 km; 1 month	global
		surface radiation budget components	$\pm$ 10 watts m^{-2}	500 km; 1 month	global
	parameterization of radiation & cloud feedback	spectral rad'nces of optically active constit's at solar/thermal wavelengths **	$<$ 1%	5-250 km; 1 month	selected regions
	cloud properties	(i) local cld-top ht, ht of base, thickness, microphysics, water/ice content radiation fluxes, grnd based radar	as needed for local studies	local	local studies carried out on variety of cloud types and cloud systems
		(ii) global *** amount cld-top ht. thickness radiation at top of atmosphere	0.05 0.5 to 1 km 0.5 to 1 km 1%	global	
	cross-correlation of net radiation with climate quantities	emitted LW rad'n planetary albedo surface albedo	$\pm$ 5 watts m^{-2} $\pm$ 1% $\pm$ 3%	250 km; 1 month 250 km; 1 month 250 km; 1 month	global

NOTES: * here accuracy is absolute, referred ultimately to primary basic standard; precision is relative accuracy, i.e., smallest detectable signal that is larger than instrument system random noise

** observations should be made as part of a special observing program

*** requirements to be satisfied by operational meteorological/environmental satellites

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APPENDIX B

EARTH RADIATION MODELS

INTRODUCTION

In its simplest form, the ERB experiments are intended to measure a simple quantity - the power per unit over input or exiting from the top of the atmosphere over time scales long enough for the power to have an impact on the weather or climate. This simple view of the problem suffers from the fact that the top of the earth's atmosphere is being treated as though it were a sheet metal cover to the atmosphere. However, provided that the top of the atmosphere is taken to be indicative of an altitude reasonably higher than the levels at which ozone absorption is important, the top of the atmosphere appears to be a well defined quantity. It should be noted that because a sphere with radius 6371 + 50 km has 3.2×10^6 km^2 more area than a sphere of radius 6371 + 30 km, a constant power per unit area going through the smaller sphere will result in a 0.6% smaller power per unit area going through the larger. Thus, if 240 wm^{-2} are going out through the smaller sphere, then 238.5 wm^{-2} are going out through a sphere with the top of the atmosphere at 50 km.

Aside from the difficulty in defining the top of the atmosphere, the measurement of the reflected or emitted radiant exitance suffers from a more fundamental difficulty: the radiation emerges not in one direction but in many. As a result, it is necessary to reduce the number of degrees of freedom by developing earth radiation models of spectral, angular, spatial, and temporal variations.

In quantitative work, the basic variation in angle and wavelength is described by the radiance, L_λ. If we consider a given (theoretically

miniscule) area on our (now well defined) top of the atmosphere, then the power carried away by light travelling within a given set of directions and also within a narrow wavelength band is given by

$$dE = L_\lambda \, d\lambda \, d\Omega \cos\theta \, dA,$$

as illustrated by Figure 1.

The radiant exitance is then found by adding together the contributions to the power leaving dA and dividing by the area:

$$M_\lambda = \int d\Omega \cos\theta \, L_\lambda \, .$$

Because M_λ represents a sort of average (or moment) of L_λ, it is convenient for our purposes to define a function

$$R(\underline{\Omega}) \equiv L/M$$

which describes the angular variation of the radiation field, but not its strength. For the shortwave reflected radiation, R is (to within a constant) known as the bidirectional reflectance. For the longwave radiation R is known as the limb darkening function. One of the major tasks of providing earth radiation models is to provide reasonable estimates of R.

Unfortunately, there appears to be no hope of directly measuring M at the top of the atmosphere because we would have to integrate over all directions at every point on this "surface". Because of the fact that radiance is constant along a path unless the atmosphere gets in the way, it can be shown through Gauss' Theorem that if we measure $\int dAM$ on some surface surrounding the TOA, we can simply take this power and reduce it to the TOA by taking

the radio of the areas. In this sense, the problem reduces to that of simply sampling M on the outer sphere well enough so that we can be sure we have measured the total power accurately. This is probably the basis for some of the early concepts of earth radiation budget measurement, as suggested by Figure 2.

The trouble is that as our understanding of the problem has increased, so has our desire for more detailed knowledge. We now hope to produce from measurements of the radiant exitance on the integrating sphere values of the radiant exitance as a function of position on the TOA. As we have already stated, our interest lies in pole-equator radiation balances and in longitudinal variations in M.

When we wish to quantitatively relate the measurement to M, we must return to a geometry similar to that shown in Figure 1, in which we relate the power emitted by an area dA which is received by an area dA' on the detector:

$$dE = L_\lambda d\lambda \frac{dA' \cos\theta'}{\rho^2} \cos\theta \, dA \, .$$

This relationship constitutes the fundamental "law" in analyzing the relationship between the satellite signal and the quantity we desire.

Not all areas on the earth are dark, of course, so the total power falling on the detector involves an integration over the area of the top of the earth's atmosphere visible to the detector. The detector converts this incident power into a signal with a sensitivity $S_\lambda(\theta', \phi')$, so that the part of the output signal coming directly from the earth is

$$m = \underset{\substack{\text{visible}\\\text{earth}\\\text{area}}}{\int dA} \int d\lambda \underset{\substack{\text{detector}\\\text{area}}}{\int dA'} \; S_\lambda \left[\frac{\cos \; \cos\theta'}{\rho}\right] R_\lambda M \, .$$

Although it may not be apparent from the form of the integration, the integral over dA is equivalent to an integration over the field-of-view. Although a treatment of other influences and errors is given elsewhere in this report, this equation should be considered to be the fundamental equation relating a given measurement m to the quantity desired M. It applies to both narrow and wide field-of-view instruments. Of course, for a NFOV scanner channel, the integration over the earth's surface becomes small enough that we usually expect to be able to assume both R and M are constant across it, so that they can be removed from the integral as constants.

Although the product $R_\lambda M$ could as easily have been written L_λ (the radiance), we have used directional-spectral function R_λ because of the large number of quantities that would otherwise be required. As we have tried to make abundantly clear, our real interest is in M as it varies over the top of the atmosphere. This variation may readily be described by assuming that the earth's TOA is divided into a number of tiles of roughly equal area, e.g. 250 km x 250 km (or 2 1/2° lat x 2 1/2° long. at the equator). Table 1 shows the number of tiles required for various resolutions, as well as the number of L values that would be required to give the same resolution in direction of propagation as is required in space

Resolution	10°x10° (1000km x 1000km)	5°x5° (500km x 500km)	2 1/2° x 2 1/2° (250km x 250km)
No. of M Values	100	400	1600
No. of L values for identical resolution in M	5000	80000	1280000

TABLE 1

Although the construction of M would be possible from observations of L, the number of quantities to be determined is clearly enough to clog the computer when 2 1/2° x 2 1/2° resolution is desired, even if enough values can be sampled to perform the integration empirically.

Physics of the Angular and Spectral Variations

It is clear that the quantity R is important in relating the measurements to the radiant exitance that we desire. Our success in reducing the data is clearly going to be related to our ability to determine it and to characterize it by as few numbers as possible. In a broad view, R contains four possible types of variations that depend upon the nature of the underlying surface, as outlined in Table 2.

In order to explain briefly the data summarized in Table 2, it will be necessary to comment briefly on the nature of the interactions between the earth, its atmosphere, and the reflected or emitted radiation.

In the shortwave portion of the spectrum, from about 200nm(.2μm) to perhaps 3000-5000nm (3-5μm), the light impinging on the detector is reflected

sunlight. The solar spectrum has been studied intensively, and numerical values of the monochromatic irradiance from the sun are available from several sources (e.g. Thekekera, 1971; Labs and Leckel). Since most observations of the solar spectrum have been made at the ground, the published values tend to cluster together in the visual spectral range (roughly 0.35μm to 0.7μm) and to be somewhat more uncertain in the UV (owing to ozone absorption) and in the near IR (owing to water vapor absorption).

Over clear scenes, the sunlight is reflected from the ground and scattered by the atmosphere. Ozone absorbs almost all incident sunlight with wavelengths shorter than about 0.35 μm. In unpolluted air, the light scattered from the atmosphere has been deflected by Rayleigh scattering from molecules, so that the reflected light is inversely proportional to the fourth power of the wavelength. In addition, because the scattering occurs in an optically thin atmosphere, the reflected light is strongly limp brightened. To a fair approximation, the radiation emerging has a radiance distribution that is approximately given by

$$L_\lambda = M_{\lambda o} \left[\frac{\tau_o}{4\pi\cos\theta} \left(\frac{\lambda_o}{\lambda}\right)^4 + a_{surf}\right] .$$

Atmospheric haze will tend to decrease both the wavelength dependence and angular variation in the reflected radiance. The major complicating factor is the surface reflection, which constitutes about one-half of the reflected solar radiation. Over the ocean, the reflection may assume a nearly specular character, whose half-width depends directly on wind speed (Kattawar, Plass, and Guinn).

Completely overcast scenes or snow, on the other hand, have reflection that is dominated by multiple scattering from nearly nonabsorbing particles. As a result, except for absorption by water vapor in the near IR bands, the reflection is not strongly dependent on wavelength. The angular variation in this situation is more difficult to characterize, although it may be treated theoretically - albeit with considerable caution owing to the strongly forward peaked scattering functions of large particles. General references include Pattridge and Platt, 1976; Travis and Hansen, 1975; Irvine. Roughly speaking, when the sun is overhead, the reflected radiation should be reasonably close to Lambertian. As the sun moves towards the horizon, the reflected radiation tends to become more strongly forward peaked since the light is reflected after one or two scatterings and still carries much information about its interactions with the water droplets or ice crystals making up the scattering medium.

Between clear conditions and completely overcast or snow lies a difficult-to-characterize area of broken or scattered cloudiness (or patchy snow). Although the reflected radiation might hopefully be characterized as composite between clear and cloudy, the situation is likely to be considerably more complicated. Theoretical studies of radiative transfer in the atmosphere have only recently advanced enough to consider the influence of cloud geometry upon reflection (Cox and McKee, 1975; Davies, Arduini and Barkstrom; and others). It is known that after sunlight enters clouds of finite horizontal extent it may leak from the sides. As a result, the larger clouds tend to be more reflective (Reynolds and McKee, 1978) a fact that has been verified experimentally. In addition shadows are introduced that can be seen from

some directions and not others. Because cloud cover, cloud size, and geometry change markedly with time of day, weather conditions, and season, the reflection from clouds constitutes the greatest uncertainty in characterizing the reflection function.

In the longwave portion of the spectrum, from about 5 μm to beyond 100 μm, the light impinging on the detector is light emitted by the earth, the gases in the atmosphere, or clouds. The basic features of the spectrum and the angular distribution are easy to understand if we recall that is any wavelength band, hot bodies emit more energy than cold bodies. Since the temperature of the atmosphere declines as we go from the ground to the tropopause and becomes reasonably constant thereafter, in transparent regions of the spectrum, emitted power is high. When the atmosphere becomes opaque in a given spectral band radiation emitted by the ground does not reach the detector outer space. Radiation that does emerge comes from high in the atmosphere and has a lower power than an equivalent high temperature source. [e.g. Kunde, et. al., 1960]

Because a light ray propagating obliquely through the atmosphere encounters more material than does radiation travelling vertically, radiation that arrives at a satellite detector after travelling obliquely through the atmosphere started higher than did normally propagating radiation. As a result, less radiation comes from the limb than from the nadii-the atmosphere appears to be neatly universally limb-darkened.

The major effect of clouds is to raise the lower boundary of the atmosphere, or to make the "ground" seem colder. This fact is routinely used to detect clouds in temperature retreivals from satellites. Thus, the

spectral variation in the longwave region is reduced, but the angular variation is not expected to be strongly affected.

In short, the physics of reflection and emission has long been studied. The major uncertainties in our understanding lie in the areas covered by broken clouds. Unfortunately, it is difficult to overcome this lack of understanding by making observations as we shall see.